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A STUDY OF THE HYDROLOGY OF EASTERN NORTH AMERICA USING ATMOSPHERIC VAPOR FLUX DATA

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ABSTRACT

The atmospheric water vapor flux divergence and certain aspects of the water balance of Eastern North America are investigated, using data from the period May 1, 1958, to Apr. 30, 1963.

Mean monthly values of evapotranspiration and storage change are computed as residuals, using measured values of vapor flux divergence, precipitation, and streamflow. Computations are performed for regions varying in size from 42×10^5 km^2 to approximately 5×10^5 km^2. The results for the smaller areas, which are the least reliable, are critically examined.

Computed values of evapotranspiration and storage change are compared with the climatological estimates of Thornthwaite Associates and Budyko. The Thornthwaite climatic water balance data appear to overestimate $\overline{P-E}$, the difference between precipitation and evapotranspiration, during winter and underestimate it during summer. Budyko's values of evapotranspiration generally show a slightly smaller seasonal variation and appear to lead the values obtained from the atmospheric budget computations by around 0.5 to 1 mo.

Flux divergence computations are made for the Gulf of Mexico and Caribbean Sea, and the results are compared with values obtained by Hastenrath and with estimates of $\overline{E-P}$ by Wüst and Budyko.

Interannual variations in storage over the Eastern Region of North America are examined and are found to be comparable with the seasonal changes. The onset of the drought of the early and mid-1960s is clearly reflected in the computed storage values.

It is found that variations in mean monthly precipitation during winter are positively correlated with the strength of the northward flow of moisture across the Gulf Coast, but little or no relationship between these quantities appears to exist during summer.

CONTENTS

1. INTRODUCTION

Consideration of the continents, oceans, and atmosphere as parts of a single interacting system is fundamental to a basic understanding of the hydrologic cycle. Consequently, the recent general circulations model experiments of Manabe et al. (1965) and Manabe (1969), which for the first time incorporated a global hydrologic cycle, have contributed significantly to a better understanding of the role of water in the earth-atmosphere system. Observational studies, whose goal is a description of the global hydrologic cycle, should be pursued simultaneously with the model studies. Unfortunately, existing observations of hydrologic parameters are not adequate to allow very detailed studies on a global or even hemispheric basis. However, data are adequate over several fairly large regions to provide a broad-scale description of some aspects of the hydrologic cycle of the particular region.

One such region is the North American sector. A number of large-scale hydrologic studies of all or portions of this region have been conducted during the past 15 yr. The first study of this type was made by Benton and Estoque (1954) [2] who attempted an evaluation of the surface water balance of the continent, using observed values of atmospheric moisture transport. More recent investigations include those of Barry (1967) over northeastern North America; Hastenrath (1966) over the Gulf of Mexico and Caribbean Sea; and on a somewhat smaller scale, Rasmussen's (1968) study of the Upper Colorado River Basin. A study of the entire North American sector has been made by the author (Rasmusson 1967, 1968, hereafter referred to as R1 and R2), and the results illustrated and discussed in this paper represent a continuation of that investigation. Water balance computations over eastern North America for basins an order of magnitude smaller than those described in R2 will be discussed, and the results for the various basins compared.[3] In addition, the results obtained from a 2-yr balance computation for the Gulf of Mexico and Caribbean Sea will be critically examined. Interannual variations in storage, vapor flux, precipitation, and vapor flux diver-

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[2] See also Benton et al. (1953).

[3] Additional discussion of these data may be found in Malhatra (1969) and Lee (1970).

gence over Eastern North America will be examined, and relationships between these quantities and variations in flux and flux divergence over the Gulf of Mexico and Caribbean Sea will be described.

The water balance equations presented and discussed in R2 may be written as

$$\left\langle\frac{\partial S}{\partial t}\right\rangle=\langle\nabla\cdot\overline{\mathbf{Q}}\rangle+\left\langle\frac{\partial W}{\partial t}\right\rangle-\langle\overline{R_0}\rangle \tag{1}$$

and

$$\langle\overline{E}\rangle=-\langle\nabla\cdot\overline{\mathbf{Q}}\rangle+\langle\overline{P}\rangle+\left\langle\frac{\partial W}{\partial t}\right\rangle. \tag{2}$$

The following notation is that of R1 and R2.

W	total water vapor content of the atmospheric column
R_0	streamflow runoff
P	precipitation
E	evapotranspiration
S	total surface and subsurface storage above a given datum
$\mathbf{Q}$	total vertically integrated flux of atmospheric water vapor
Q_λ	zonal component of $\mathbf{Q}$, positive eastward
Q_ϕ	meridional component of $\mathbf{Q}$, positive northward
$\langle(\ )\rangle$	areal mean value
$\overline{(\ )}$	time mean value

$\langle\nabla\cdot\overline{\mathbf{Q}}\rangle$, the atmospheric vapor flux divergence, and $\langle\overline{\partial W/\partial t}\rangle$ are computed from aerological observations of wind and humidity; $\langle\overline{R_0}\rangle$ is estimated from streamflow measurements; $\langle\overline{P}\rangle$ is estimated from precipitation measurements; and $\langle\overline{\partial S/\partial t}\rangle$ and $\langle\overline{E}\rangle$ are evaluated as residuals.

Measured values of precipitation used in these studies almost certainly represent systematic underestimates of the actual precipitation. This bias is in part due to wind action that reduces the catch of precipitation in gages elevated above the ground (Rodda 1967, Weiss and Wilson 1958, and Struzer et al. 1965). As would be expected from typical wind profiles, the loss of catch increases with increasing gage elevation (Bruce and Rodgers 1962). The loss is also significantly greater for snow than for rain. Consequently, seasonal variation in the character of precipitation and in the wind regime during periods of precipitation may introduce seasonal variations in the deficiency of the catch. In a comparison of a ground-level gage and a standard British rain gage at a height of 1 ft, Rodda (1967) found that the ground-level gage caught 6.6 percent more rain (8 percent more total precipitation) than did the standard gage. When considering rain only, the difference in catch was still significantly greater in winter than in summer. Somewhat greater differences can be expected over the United States, where the height of the standard gage is 31 in. and where gage distribution over mountain areas leads to a negative bias (Rasmussen 1968). Rodda suggests that the difference may be on the order of 10 percent; and in R2, an underestimation of 5–10 percent was suggested.

Mean annual values of $\overline{E}$, whether computed from a terrestrial water balance between $\overline{R_0}$, $\overline{P}$, and $\overline{E}$ or an atmospheric balance between $\nabla\cdot\overline{\mathbf{Q}}$, $\overline{P}$, and $\overline{E}$, will be biased to the same extent as the values of $\overline{P}$. On the other hand, the difference between $\overline{P}$ and $\overline{E}$ can be computed from the atmospheric vapor balance without facing the serious problems involved in the determination of $\overline{P}$ and $\overline{E}$ individually.

Data sources and analysis procedures used in this study have been reviewed in R1 and R2. Discussions on the reliability and uncertainty of the atmospheric vapor flux data are also included in these papers. Values of flux divergence over the North American Continent are for the 5-yr period May 1958–April 1963 and were obtained from the computer-analyzed maps of Bock et al. (1966). Values of flux divergence over the Central American Sea are from the 2-yr period May 1961–April 1963 and were obtained from hand-analyzed maps (Rasmusson 1966). For a general discussion of the balance equations and the assumptions involved in the computations, see R2.

2. MEAN CONDITIONS—EASTERN NORTH AMERICA

Measured values of $\langle\overline{R_0}\rangle$ and $\langle\overline{P}\rangle$ and computed values of $\langle\overline{E}\rangle$ and $\langle S\rangle$ for the combined Central Plains and Eastern Region (fig. 1) were presented in R2. Mean values for the two individual areas are given in tables 1 and 2 and illustrated in figures 2 and 3. Averages for the Central Plains are for the 5-yr period May 1958–April 1963. Abnormally dry conditions existed over large portions of the East during the final year of this 5-yr period. For this reason, averages for the Eastern Region were computed from only the first 4 yr of the period.

As in R2, it was assumed that any computed net storage change from beginning to end of the averaging period was due to constant systematic error in the evaluation of the vapor flux divergence. Mean monthly values of flux divergence were therefore adjusted to reduce the net storage change during the averaging period to zero. The required adjustments were -0.54 cm mo^{-1} and $+0.87$ cm mo^{-1} for the Eastern and Central Plains Regions, respectively. These values may be compared with the adjustment of $+0.35$ cm mo^{-1} required for the combined area 5-yr average.

The Central Plains Region consists, for the most part, of the relatively dry low-runoff regions of the Great Plains. Its average annual rainfall of 60.8 cm was less than 60 percent of that for the Eastern Region. Streamflow constitutes a relatively minor component of the hydrologic cycle, amounting to only 16 percent of the precipitation and 19 percent of the mean annual evapotranspiration. In comparison, mean annual streamflow from the Eastern Region amounted to 40.3 cm, 39 percent of annual precipitation and 63 percent of annual evapotranspiration.

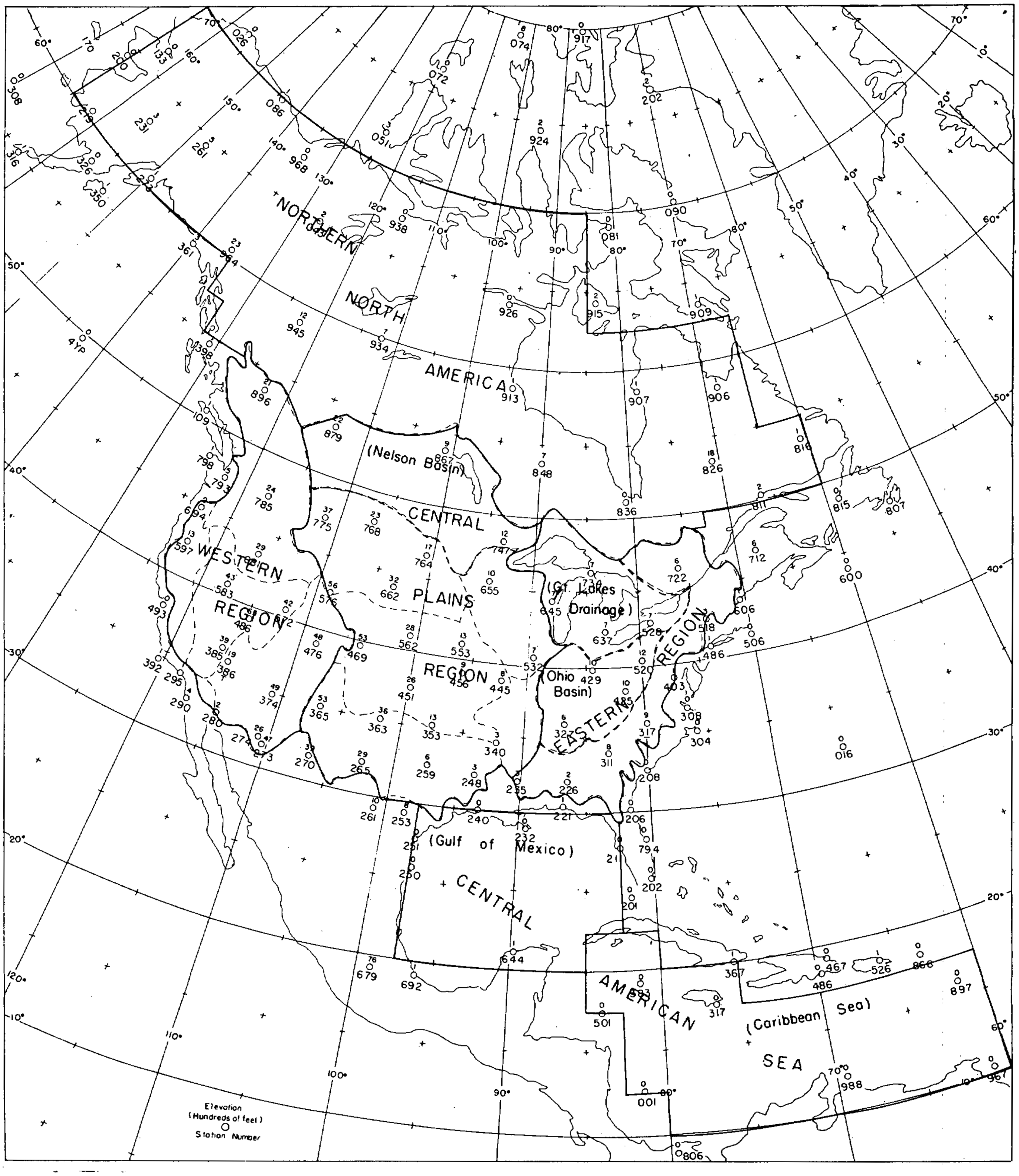

FIGURE 1.—Regions of water balance computations (from Rasmusson 1968).

The Great Lakes account for more than 10 percent of the area of the Eastern Region, and the mean monthly values of storage are significantly influenced by the presence of the lakes. The seasonal storage in the lakes follows a pattern that is almost out of phase with the remainder of the region (fig. 5), lake storage being highest in midsummer and lowest in late winter. Values of $\langle\overline{E}\rangle$ are also influenced by the presence of the lakes (fig. 5),

TABLE 1.—*Central Plains Region (area=42×10⁵ km²) computed mean monthly water balance components, May 1958–April 1963*

Month	$\langle \overline{R}_0 \rangle$ (cm mo⁻¹)	$\langle \overline{P} \rangle$ (cm mo⁻¹)	$\langle \overline{E} \rangle$ (cm mo⁻¹)	$\langle \Delta S \rangle$ (cm mo⁻¹)	$\langle S \rangle$* (cm)
September	0.57	6.60	5.02	+1.01	1.01
October	.68	4.63	3.51	+.44	1.45
November	.53	3.42	2.37	+.52	1.97
December	.59	3.02	1.78	+.65	2.62
January	.61	2.33	1.79	−.07	2.55
February	.65	2.93	1.15	+1.13	3.68
March	.87	3.82	1.79	+1.16	4.84
April	1.20	4.25	3.52	−.47	4.37
May	1.28	7.63	6.03	+.32	4.69
June	1.03	7.97	7.76	−.82	3.87
July	.89	8.18	9.41	−2.12	1.75
August	.71	6.05	7.09	−1.75	.00
Total	9.6	60.8	51.2		

*Storage on the last day of the month minus that on August 31

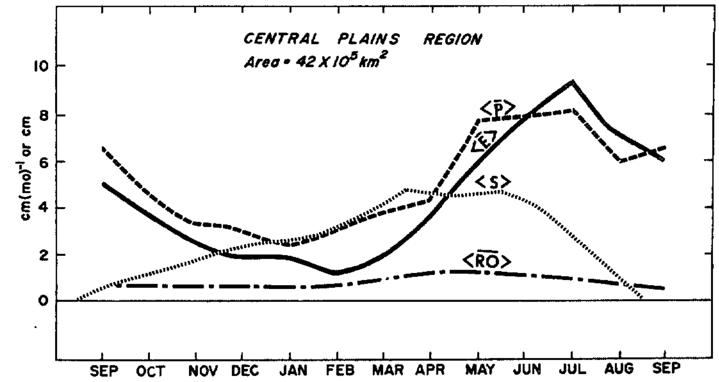

FIGURE 2.—Computed water balance components for the Central Plains Region.

TABLE 2.—*Eastern Region (area=22×10⁵ km²) computed mean monthly water balance components, May 1958–April 1962*

Month	$\langle \overline{R}_0 \rangle$ (cm mo⁻¹)	$\langle \overline{P} \rangle$ (cm mo⁻¹)	$\langle \overline{E} \rangle$ (cm mo⁻¹)	$\langle \Delta S \rangle$ (cm mo⁻¹)	$\langle S \rangle$* (cm)
September	1.96	8.52	5.31	1.26	1.26
October	2.02	7.62	4.98	.62	1.88
November	2.16	7.32	4.05	1.12	3.00
December	3.07	7.32	3.24	1.03	4.03
January	3.49	7.60	3.54	.58	4.61
February	4.07	8.88	1.42	3.38	7.99
March	5.51	8.02	3.42	−.92	7.07
April	5.70	8.58	4.98	−2.12	4.95
May	4.51	8.92	7.01	−2.61	2.34
June	2.86	10.40	9.00	−1.46	.88
July	2.63	11.28	8.82	−.18	.70
August	2.31	9.72	8.17	−.70	.00
Total	40.3	104.2	63.9	0.0	

*Storage on the last day of the month minus that on August 31

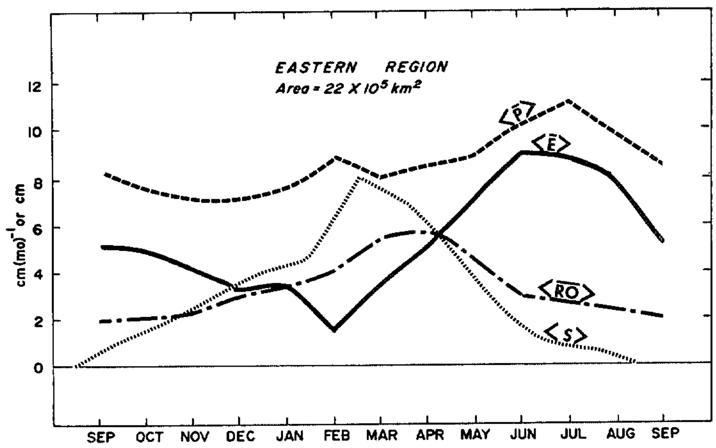

FIGURE 3.—Computed water balance components for the Eastern Region.

TABLE 3.—*Truncated Eastern Region* (area=19.5×10⁵ km²) computed mean monthly water balance components, May 1958–April 1962*

Month	$\langle \overline{R}_0 \rangle$ (cm mo⁻¹)	$\langle \overline{P} \rangle$ (cm mo⁻¹)	$\langle \overline{E} \rangle$ (cm mo⁻¹)	$\langle \Delta S \rangle$ (cm mo⁻¹)	$\langle S \rangle$† (cm)
September	1.81	8.40	4.62	1.97	1.97
October	1.98	7.81	4.23	1.63	3.60
November	2.31	7.40	3.11	1.97	5.57
December	3.31	7.54	1.84	2.39	7.96
January	3.75	7.87	3.30	.82	8.78
February	4.49	9.40	.88	4.01	12.79
March	6.45	8.45	3.49	−1.49	11.30
April	7.17	8.70	5.47	−3.94	7.36
May	5.35	9.07	7.91	−4.19	3.17
June	2.90	10.80	9.92	−2.02	1.15
July	2.59	11.75	9.60	−.44	.71
August	2.09	9.85	8.47	−.71	.00
Total	44.2	107.0	62.8	0.0	

*Averages for the Eastern Region (exclusive of the Great Lakes) are based on the period May 1959–April 1963. The averages for the Great Lakes are based on the period May 1958–April 1963.

† Storage on the last day of the month minus that on August 31

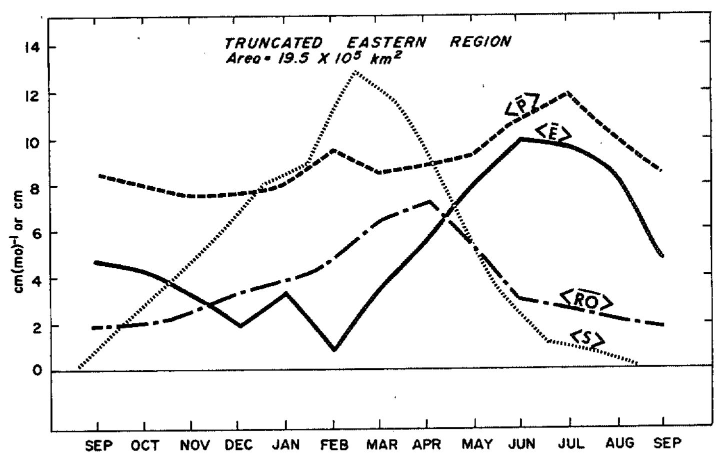

FIGURE 4.—Computed water balance components for the Truncated Eastern Region.

which have a fall and early winter maximum and a late spring and early summer minimum. It is therefore desirable to obtain a water balance for a "Truncated Eastern Region," defined as the Eastern Region less the area occupied by the surface of the five Great Lakes. This was accomplished by computing a terrestrial water balance for the lakes (see section 3). Mean monthly values of $\langle \overline{R}_0 \rangle$, $\langle \overline{P} \rangle$, $\langle \overline{E} \rangle$, and $\langle \Delta S \rangle$ for the Truncated Eastern Region

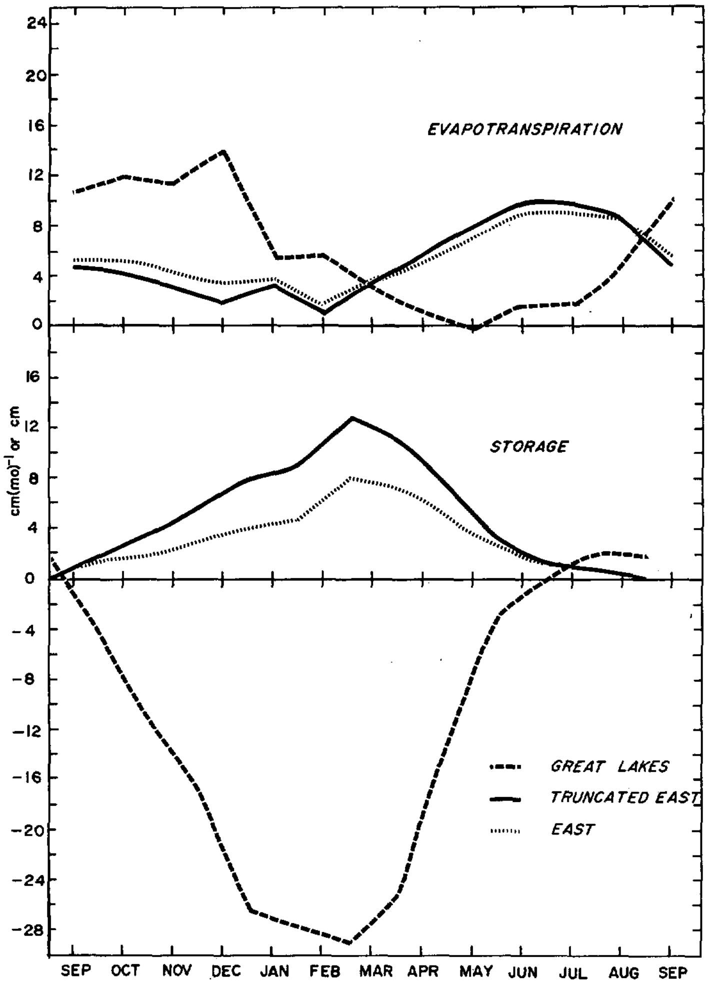

FIGURE 5.—Comparison of computed storage and evapotranspiration for the Great Lakes, Eastern Region, and Truncated Eastern Region.

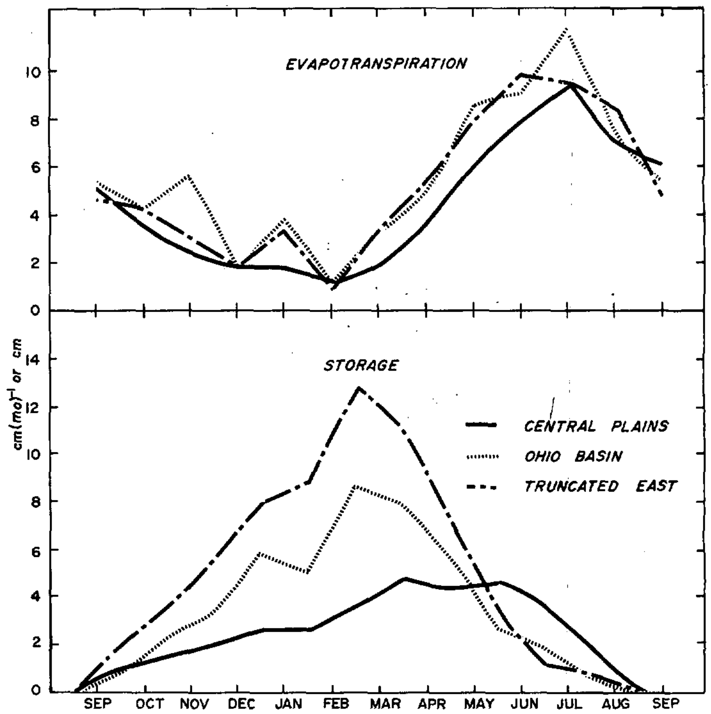

FIGURE 6.—Comparison of computed storage and evapotranspiration for the Central Plains, Ohio Basin, and Truncated Eastern Region.

are given in table 3 and shown on figure 4.

The computed loss in storage during spring and summer (March 1 to September 1) over the Truncated Eastern Region was 13.8 cm. This, together with the precipitation during the period (58.6 cm) must balance the losses due to runoff (26.6 cm) and evapotranspiration (44.8 cm). Consequently, around 20 percent of the spring and summer losses are supplied from storage accumulated during the winter. For the period April 1–July 1, computations indicate that storage provided more than 25 percent of the losses.

The amplitude of the computed storage curve for the Central Plains is around one-third that for the Truncated East (fig. 6). Computed storage losses were primarily confined to June, July, and August, during which time storage supplied 17.5 percent of the total streamflow and evapotranspiration losses.

$\langle \overline{E} \rangle$ over the Truncated Eastern Region reached a maximum of 9.9 cm in June. However, inclusion of the fifth year in the averaging period would have shifted the maximum to July. Average evapotranspiration during the period November–March was around 2½ cm mo^{-1}. Little significance should be attached to the relatively high value in January and low value in February. Such irregularities may be due to the relatively short period of 4 yr over which the averages are being computed.

Computed mean annual evapotranspiration over the Truncated Eastern Region (63 cm) was, as would be expected, considerably higher than that computed for the Central Plains (51 cm). Differences between the two areas were largest during winter and smallest during midsummer (fig. 6). $\langle \overline{E} \rangle$ slightly exceeded $\langle \overline{P} \rangle$ during July and August over the Central Plains, while $\langle \overline{P} \rangle$ always exceeded $\langle \overline{E} \rangle$ over the East.

We may, as in R2, compare the values of $\langle \overline{E} \rangle$ and $\langle S \rangle$ computed from the atmospheric water budget with those obtained by an analysis of station values obtained from the publications of Thornthwaite Associates (1964*a*, 1964*b*). The comparisons are shown in table 4 and on figures 7 and 8. Mean annual values of evapotranspiration are in good agreement; but as in the case of the larger areas described in R2, the Thornthwaite values are smaller in winter and larger during summer. This leads to a storage curve of significantly greater amplitude than that computed from the atmospheric water budget.

These Thornthwaite storage values, which have previously been used for large-scale water balance computations by a number of investigators, were apparently not meant to be strictly comparable to the total storage change over a large basin. According to van Hylckama (1956), the Thornthwaite computational procedure con-

TABLE 4.—*Comparison of storage and evapotranspiration estimates; units, cm or cm mo*$^{-1}$

Month	Central Plains Region					Truncated Eastern Region				
	$\langle \overline{E} \rangle$			$\langle S \rangle$		$\langle \overline{E} \rangle$			$\langle S \rangle$	
September	5.1*	4.5†	6.2‡	1.0*	0.0‡	4.6*	5.5†	8.4‡	2.0*	0.0‡
October	3.5	3.0	3.8	1.5	.6	4.2	4.0	4.6	3.6	2.1
November	2.4	2.0	.9	2.0	3.0	3.1	2.5	1.3	5.6	7.1
December	1.8	1.2	.2	2.6	5.5	1.8	1.7	.3	8.0	12.5
January	1.8	1.0	.2	2.6	7.3	3.3	1.5	.3	8.8	16.5
February	1.2	1.7	.3	3.7	8.8	.9	2.4	.4	12.8	19.2
March	1.8	3.0	1.0	4.8	9.7	3.5	4.2	1.3	11.3	18.6
April	3.5	4.5	3.8	4.4	9.6	5.5	6.3	3.9	7.4	10.3
May	6.0	6.7	7.6	4.7	8.9	7.9	8.0	8.6	3.2	9.3
June	7.8	7.8	10.4	3.9	6.8	9.9	8.7	12.2	1.2	6.4
July	9.4	7.4	10.5	1.8	3.0	9.6	8.5	13.5	.7	2.8
August	7.1	6.0	8.7	.0	.6	8.5	6.7	11.5	.0	.1
Total	51.2	48.8	53.6			62.8	60.0	66.3		

*Figures in this column are from the atmospheric water balance estimate.
†Figures in this column are from the Budyko evapotranspiration estimate.
‡Figures in this column are from the Thornthwaite climatological estimates.

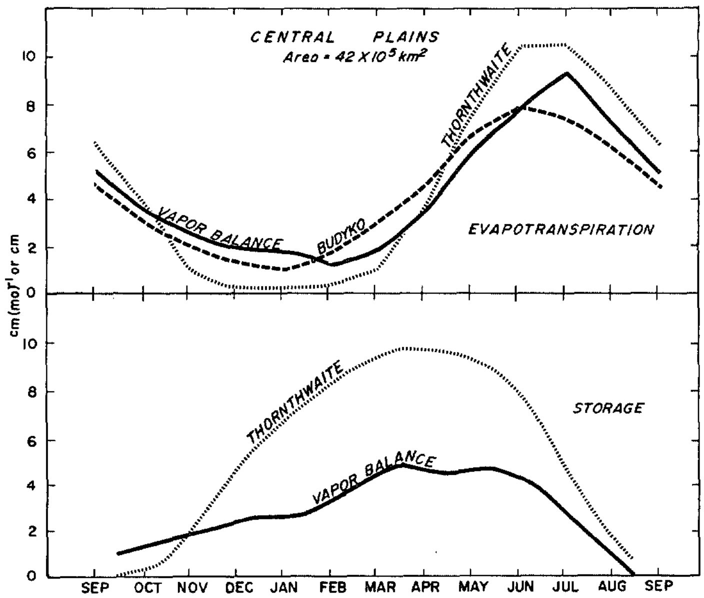

FIGURE 7.—Comparison of estimates of evapotranspiration and storage for the Central Plains Region.

siders water to go from storage to runoff where it first reaches the local lake or stream. Thus his storage figures do not account for any storage changes that may occur in the lakes and streams within the basin. Such storage changes may not be negligible in basins where lake and channel storage are significant, as may be the case for the area draining into the Great Lakes.

Values of $\langle \overline{E} \rangle$ estimated from the maps of Budyko (1963) are also compared with those obtained from the atmospheric vapor balance (figs. 7 and 8). The seasonal variation in Budyko's values appears to exhibit a slightly smaller amplitude and to lead the values obtained from the atmospheric vapor balance computations by around 0.5 to 1 mo.

3. BALANCE COMPUTATIONS—OHIO BASIN AND GREAT LAKES DRAINAGE

For a given aerological network and averaging period, the probability of an accurate evaluation of vapor flux divergence decreases as the size of the area decreases. This decrease in accuracy occurs for a number of reasons. Most obvious is the fact that changes in the area over which averages are taken are proportional to changes in L^2 (where L is a typical length), while changes in the perimeter along which the inflow and outflow are evaluated are proportional to $4L$. To retain the same degree of accuracy in the divergence computations as the area is decreased, one is faced with the necessity of more accurately evaluating the differences between inflow and outflow from the area.

Another source of error arises from the inability of the aerological network to resolve the small-scale large amplitude features of the divergence field. Since mean annual runoff will very nearly equal the mean annual value of $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$ over most areas, one can readily establish the importance of these small-scale features by examining figure

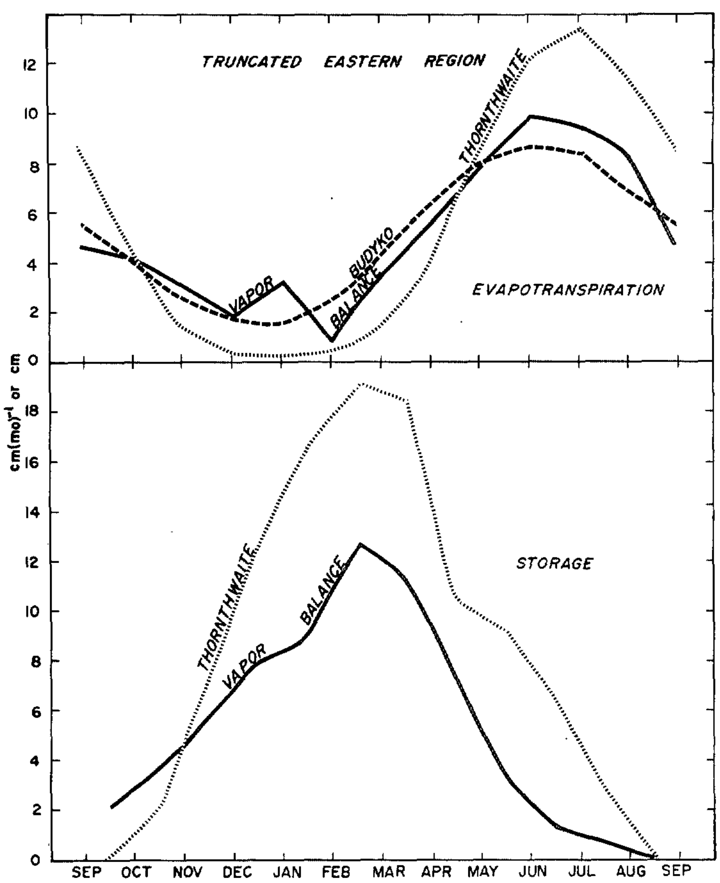

FIGURE 8.—Comparison of estimates of evapotranspiration and storage for the Truncated Eastern Region.

9 which shows the main features of the mean annual runoff from the United States. Note that values on this figure are inches per year.

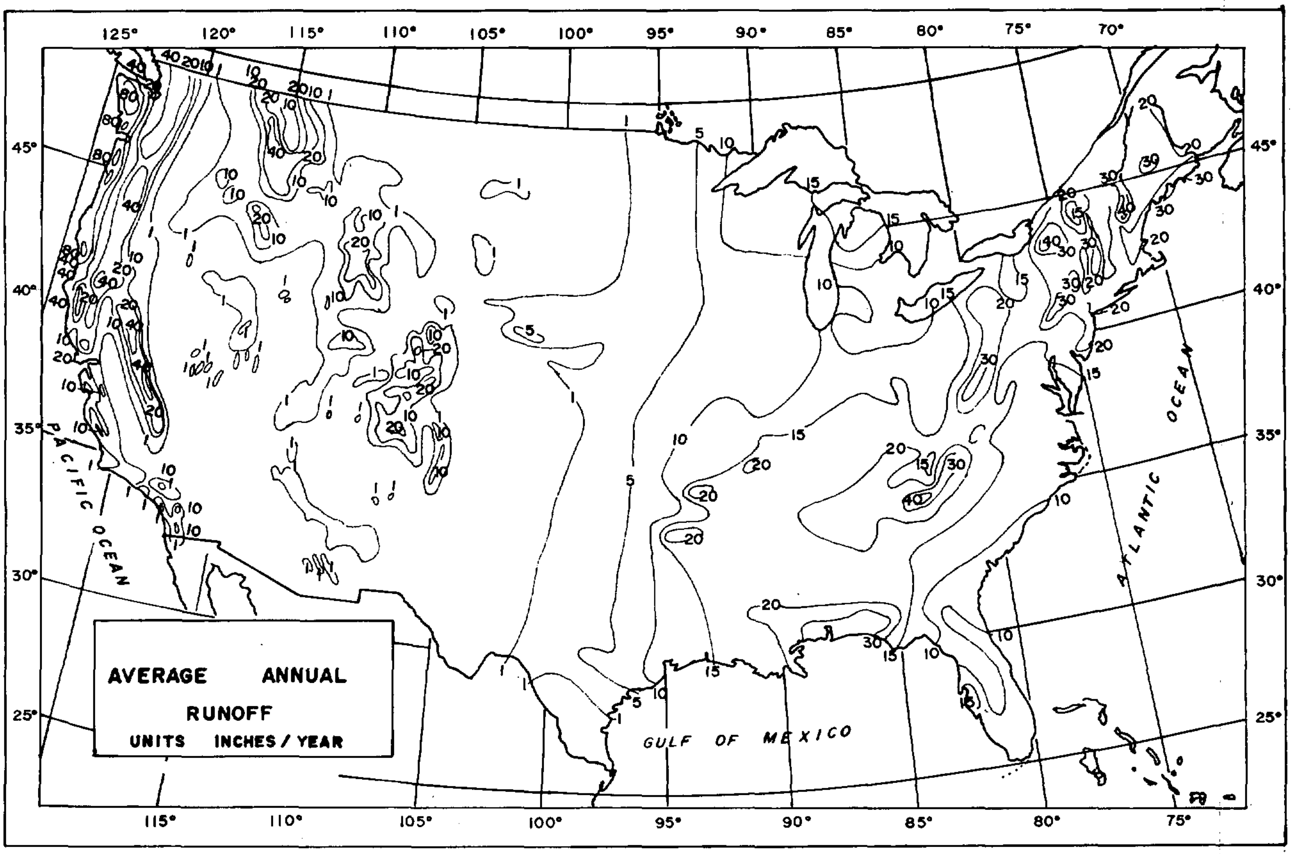

FIGURE 9.—Average annual runoff (adapted from McGuinness 1964).

Runoff variations of 25 cm yr^{-1} or more over distances of 100 km are not uncommon, and a substantial fraction of the spatial variance in the annual runoff pattern is associated with features whose dimensions range between 200–600 km. Since the typical spacing of aerological stations during the 5-yr period analyzed was generally 250–350 km, it was not possible to properly resolve such small-scale features. The situation is aggravated by the tendency for the smaller scale features to be found in the vicinity of important drainage divides, while the aerological stations are, as a rule, located in valleys. Similar small-scale features are found on individual monthly isohyetal maps and on monthly maps of departure from normal precipitation, particularly in those areas and during those seasons when a substantial amount of precipitation results from convective activity. One would therefore expect both random and systematic errors to arise from the inability to properly define these features.

Another factor that significantly increased the error over smaller areas during this 5-yr period was discussed in R2. This arose from the fact that analyses, based on twice-daily observations available during the period, exhibited a large-scale large amplitude error pattern superimposed on the real divergence field. Because of the scale of the pattern, errors tended to cancel when averaged over areas of roughly 10^6 km^2 or more. For smaller areas, particularly for those less than 5×10^5 km^2, this was often not true, leading to large errors in the evaluation of $\langle \nabla \cdot \bar{\mathbf{Q}} \rangle$. Furthermore, addition, removal, or changes in location of individual stations of the aerological network during the period of record may have an effect on the results for smaller areas.

Computations made for two areas for which individual mean monthly values appear to be unreliable but for which the 5-yr average values yield some useful information will be presented. These areas are the Ohio Basin (area$=5.2\times10^5$ km^2) and the Great Lakes Drainage above Cornwall, Ontario (area$=7.3\times10^5$ km^2).

OHIO BASIN

Results of computations for the Ohio Basin (see fig. 1) are given in table 5. The atmospheric water balance computations for this basin must be viewed with caution, due to the large divergence correction (-3.47 cm mo^{-1}) required to remove the computed 5-yr storage change. This correction is of the same order of magnitude as the divergence values themselves. It is therefore not surprising that a plot of the monthly values of $\langle \overline{E} \rangle$ and $\langle \Delta S \rangle$ (fig. 10) shows a somewhat erratic behavior. However, a relatively smooth curve results from a simple 0.25, 0.50, 0.25 smoothing of these values. A comparison of smoothed values of $\langle \overline{E} \rangle$, computed from the atmospheric water balance, with estimates of $\langle \overline{E} \rangle$ from Thornthwaite Associates (1964*a*, 1964*b*) and Budyko (1963) is given on

Table 5.—*Ohio Basin (area=5.3×10⁵ km²) computed mean monthly water balance components, May 1958–April 1963*

Month	$\langle \overline{R}_0 \rangle$ (cm mo⁻¹)	$\langle \overline{P} \rangle$ (cm mo⁻¹)	$\langle \overline{E} \rangle$ (cm mo⁻¹)	$\langle \Delta S \rangle$ (cm mo⁻¹)	$\langle S \rangle$* (cm)
September	1.33	7.61	5.48	+0.80	0.80
October	1.33	7.30	4.35	+1.60	2.40
November	1.87	8.42	5.64	+ .90	3.30
December	3.32	7.64	1.75	+2.57	5.87
January	4.95	7.83	3.85	− .99	4.88
February	4.49	9.51	1.17	+3.85	8.73
March	9.25	11.76	3.30	− .79	7.94
April	6.08	8.42	4.81	−2.47	5.47
May	5.40	11.05	8.58	−2.92	2.55
June	2.82	11.15	9.06	− .73	1.82
July	2.58	13.04	11.07	−1.29	.53
August	2.23	9.00	7.31	− .54	.00
Total	45.7	112.7	67.1		

*Storage on the last day of the month minus that on August 31

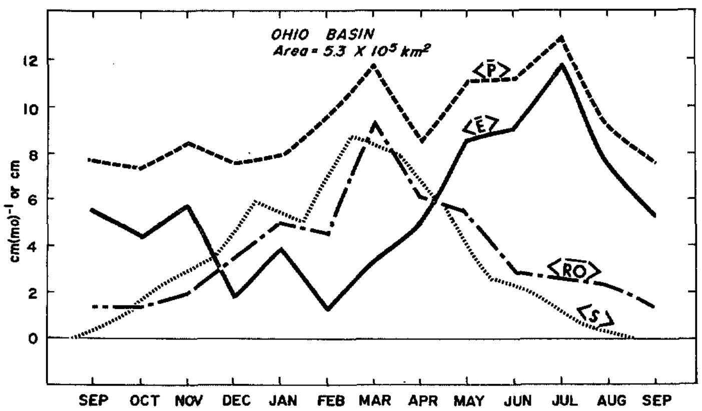

Figure 10.—Computer water balance components for the Ohio Basin.

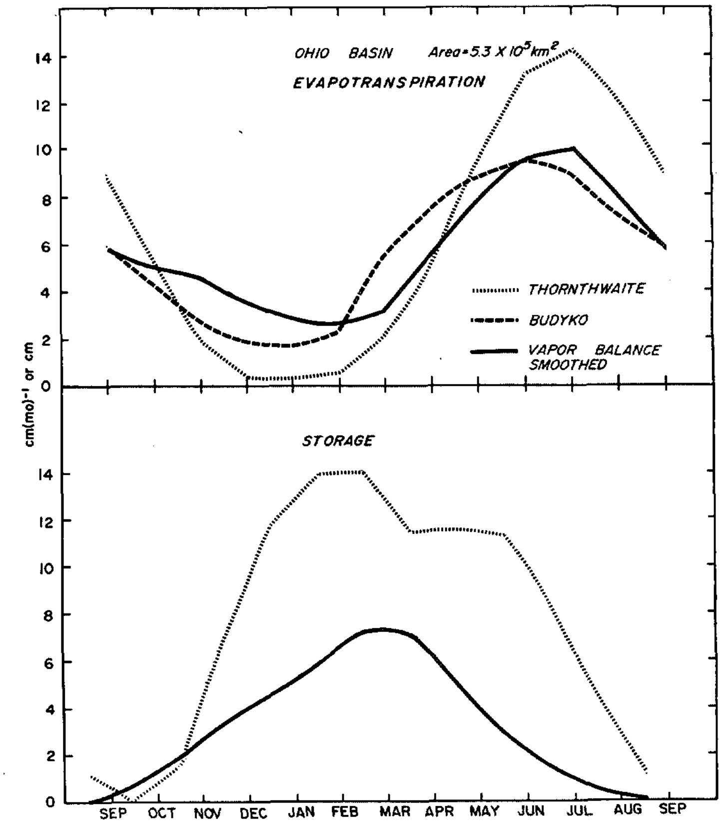

Figure 11.—Comparison of estimates of evapotranspiration and storage for the Ohio Basin.

figure 11. The smoothing operator will reduce the amplitude of the annual harmonic by about 7 percent (Holloway 1958), so that one would expect the amplitude of the annual variation to be slightly greater than that indicated. The relationship of the three estimates is similar to that previously found for the Central Plains and Eastern Regions.

GREAT LAKES DRAINAGE

Mean monthly values of precipitation for the Great Lakes and their surrounding drainage area (see fig. 1) were obtained from the U.S. Lake Survey. In this compilation, precipitation values for the lakes are estimated from shore and island measurements. For a discussion of the still unresolved controversy concerning the relationship of lake precipitation to measured shore and island precipitation, see Bruce and Rodgers (1962).

Water level gage readings from a single gage in each lake were provided by the U.S. Lake Survey, Detroit, Mich. The mean elevation on the first day of each month was used to estimate the month-to-month change in lake storage. Unfortunately, changes in elevation at a gage not only reflect changes in the mean level of the lake but also measure wind- and pressure-induced variations of the surface that are unrelated to volume changes. Use of mean daily values smooths out short-period fluctuations, but variations of synoptic time scale may still be present. It would be desirable to filter out any such variations in future studies. Fortunately, the error in estimating storage is not cumulative.

Approximately two-thirds of the total area draining into the lakes was gaged. Inflow from the ungaged areas was estimated from the flow of nearby streams. Account was taken of the diversions of water into the Lake Superior Basin and the diversion through the Chicago Sanitary and Ship Canal.

With these data, one can calculate E from the surface of the lakes from the terrestrial water balance equation

$$\langle \overline{P} \rangle + \langle \overline{I} \rangle = \langle \overline{O} \rangle + \langle \overline{E} \rangle + \langle \Delta S \rangle \tag{3}$$

where I is inflow to the lakes and O is outflow from the lakes. Mean monthly values of the parameters in eq (3), for the period May 1, 1958, to Apr. 30, 1963, are given in table 6. The parameters vary in a fairly regular manner from month to month, with $\langle \Delta S \rangle$ and $\langle \overline{E} \rangle$ being the most erratic of the five terms. Maximum values of $\langle \overline{E} \rangle$ are computed for the fall, minimum values for the spring, as ex-

TABLE 6.—*The Great Lakes (area=2.46 × 10^5 km²) computed mean monthly water balance components, May 1958–April 1963*

Month	$\langle \overline{I} \rangle$ (cm mo^{-1})	$\langle \overline{O} \rangle$ (cm mo^{-1})	$\langle \overline{P} \rangle$ (cm mo^{-1})	$\langle \overline{E} \rangle$ (cm mo^{-1})	$\langle \Delta S \rangle$ (cm mo^{-1})	$\langle S \rangle$* (cm)
September	3.22	6.35	9.28	10.43	−4.28	−4.28
October	3.95	6.23	6.95	11.85	−7.18	−11.46
November	4.92	5.95	6.55	11.02	−5.50	−16.96
December	4.90	6.08	5.49	13.90	−9.59	−26.54
January	4.78	5.92	5.22	5.19	−1.11	−27.65
February	4.65	5.37	4.71	5.45	−1.46	−29.11
March	7.95	5.95	4.51	2.81	3.70	−25.41
April	12.10	6.08	7.39	1.00	12.41	−13.00
May	9.22	6.95	7.49	−.28	10.04	−2.94
June	4.62	7.10	7.11	1.54	3.09	.13
July	3.42	7.21	7.39	1.59	2.01	2.14
August	2.84	6.88	8.51	4.94	−.47	1.67
Total	66.6	76.1	80.6	69.5		

*Storage on the last day of the month minus that on August 31. The nonzero value shown for August arises from the fact that the lakes showed an average annual storage change of +1.67 during this 5-yr period.

TABLE 7.—*Great Lakes Basin (area=7.3×10^5 km²) computed mean monthly water balance components, May 1958–April 1963*

Month	$\langle \overline{R}_0 \rangle$ (cm mo^{-1})	$\langle \overline{P} \rangle$ (cm mo^{-1})	$\langle \overline{E} \rangle$ (cm mo^{-1})	$\langle \Delta S \rangle$ (cm mo^{-1})	$\langle S \rangle$* (cm)
September	2.14	9.06	4.93	1.99	1.99
October	2.10	6.86	3.89	.87	2.86
November	2.01	6.33	1.80	2.52	5.38
December	2.05	5.07	2.50	.52	5.90
January	1.99	4.93	2.76	.18	6.08
February	1.82	4.77	1.54	1.41	7.49
March	2.02	4.55	1.44	1.09	8.58
April	2.05	6.91	2.88	1.98	10.56
May	2.34	7.60	5.17	.09	10.65
June	2.39	7.54	8.83	−3.68	6.97
July	2.43	7.97	8.80	−3.26	3.71
August	2.32	8.74	9.57	−3.15	.56
Total	25.7	80.3	54.1		

*Storage on the last day of the month minus that on August 31. The nonzero value shown for August arises from the average annual storage change in the lakes themselves during the 5-yr period.

pected. The small negative value computed for May indicates a net condensation of moisture on the relatively cold water surface. From March through May, the variation in the computed values of $\langle \overline{E} \rangle$ represent more or less a balance between $\langle \overline{P} \rangle$, $\langle \overline{I} \rangle$, and $\langle \Delta S \rangle$, while from August through February the variations are largely determined by the estimates of $\langle \Delta S \rangle$. Thus, such features as the decrease in $\langle \overline{E} \rangle$ in November and the sharp increase in December, reflected in reverse in $\langle \Delta S \rangle$, may be partly the result of inaccuracy in the estimation of the lake storage on the first day of the month.

Storage in the lakes increased during the first half of the 5-yr period and decreased during the second half, with an average change of +1.67 cm yr^{-1}.

Water balance computations for the entire basin, using the atmospheric vapor balance for the evaluation of $\langle \overline{E} \rangle$ and $\langle \Delta S \rangle$ are given in table 7 and shown on figure 12. The divergence adjustment required to reduce the net 5-yr storage change to that observed in the lakes themselves was a surprisingly small 0.03 cm mo^{-1}. Mean annual runoff from the basin is approximately 27 cm (Bruce and Rodgers 1962), mean annual precipitation approximately 79–80 cm (Richards 1965), and mean annual evapotranspiration 52 cm. Average annual values computed for the 5-yr period May 1958–April 1963 were 25.7 cm for runoff, 80.3 cm for precipitation, and 54.1 cm for evapotranspiration. Thus mean conditions during the 5-yr period were near the long-term normals.

Using the values in tables 6 and 7, one can construct a balance for the area draining into the lakes. These values are given in table 8 and on figure 13. Although the general shapes of the $\langle \overline{E} \rangle$ and $\langle S \rangle$ curves are more or less as expected, they cannot be taken at face value quantitatively. The most disturbing aspect of the results is the tendency for negative values of $\langle \overline{E} \rangle$ to appear during the fall and winter months. It is indeed hazardous to attempt an

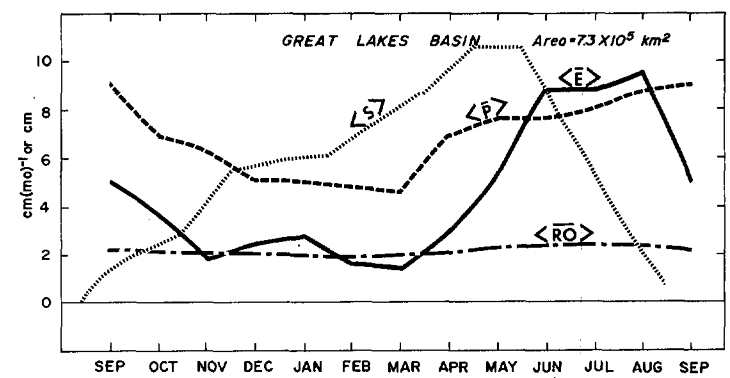

FIGURE 12.—Computed water balance components for the Great Lakes Basin.

explanation of this obvious inaccuracy without further detailed investigation. However, careful examination of the values in tables 6–8 for possible sources of error large enough to account for the apparent error in $\langle \overline{E} \rangle$ suggests the most likely sources to be the evaluation of $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$ and $\langle \overline{P} \rangle$. Adjusted mean annual values of $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$ are in error by the amount of mean annual storage change over the drainage basin during the 5-yr period. This is probably a relatively small quantity. Thus an undercomputation of $\langle \nabla \cdot \overline{\mathbf{Q}} \rangle$ during the fall and winter months, which would result in an undercomputation of $\langle \overline{E} \rangle$, will have to be compensated by an overcomputation during the spring and summer. This in turn would lead to a storage curve of excessive amplitude. Comparison of the results for this basin with those computed for the East, Central Plains, and Ohio strongly suggest this to be the case.

Underestimation of $\langle \overline{P} \rangle$ and consequent underestimation of $\langle \overline{E} \rangle$ would probably be most pronounced during the colder and more stormy period when the negative values of $\langle \overline{E} \rangle$ are computed. For example, average observed precipitation during the period October–March

TABLE 8.—*Great Lakes Basin exclusive of Great Lakes (area=4.8 $\times 10^5$ km²) computed mean monthly water balance components*

Month	$\langle \overline{R_0} \rangle$ (cm mo⁻¹)	$\langle \overline{P} \rangle$ (cm mo⁻¹)	$\langle \overline{E} \rangle$ (cm mo⁻¹)	$\langle \Delta S \rangle$ (cm mo⁻¹)	$\langle S \rangle$* (cm)
September	1.64	8.96	2.15	5.17	5.17
October	2.02	6.83	−.15	4.96	10.13
November	2.50	6.24	−2.85	6.59	16.72
December	2.49	4.87	−3.17	5.65	22.37
January	2.43	4.82	1.56	.83	23.20
February	2.36	4.83	−.41	2.86	26.06
March	4.04	4.58	.77	−.23	25.83
April	6.15	6.72	3.87	−3.30	22.53
May	4.68	7.66	7.93	−4.95	17.58
June	2.35	7.66	12.41	−7.10	10.48
July	1.75	8.25	12.45	−5.95	4.53
August	1.44	8.88	11.95	−4.51	.00
Total	33.9	80.3	46.5		

*Storage on the last day of the month minus that on August 31

was 5.4 cm mo⁻¹, while computed evapotranspiration was −0.7 cm mo⁻¹. Assuming a true value of $\langle \overline{E} \rangle$ during this period of 1.0 cm mo⁻¹ requires the actual precipitation to average 1.7 cm mo⁻¹ greater than observed. If true, this would represent an average underestimation of 24 percent for the fall and winter precipitation. Construction of a proper water balance for this area must await resolution of these uncertainties.

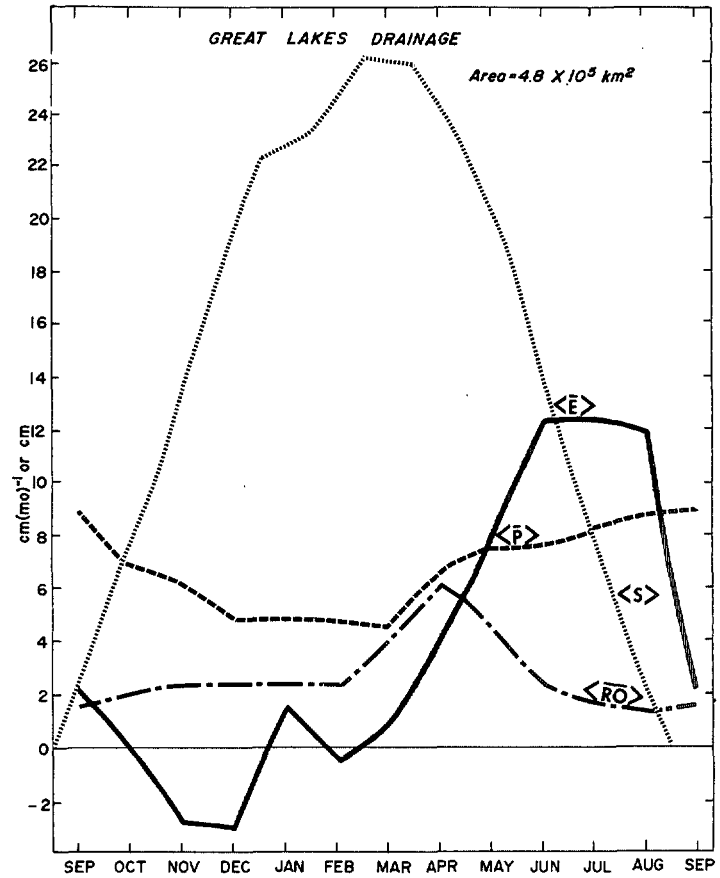

FIGURE 13.—Computed water balance components for the Great Lakes Basin exclusive of the Great Lakes.

4. BALANCE COMPUTATIONS CENTRAL AMERICAN SEA

In a discussion of the hydrology of Eastern North America, it is of interest to include a description of conditions over the Gulf of Mexico and to a lesser extent conditions over the Caribbean Sea. The characteristics of the vapor flux field over these areas have been discussed in R2. In this section, we shall review some results of water balance computations.

Over predominantly ocean areas where data are sparse, it is important that widely spaced island observations be "representative" of the large-scale atmospheric flow. Flux data obtained from island stations located some distance above sea level or strongly influenced by local surface features reflect local perturbations that are usually too small to be resolved by the observational network. Such lack of "representativeness" is particularly damaging in the case of vapor flux computations in the Tropics, for here the low-level flux is most dominant and relatively modest perturbations of the wind field give rise to significant perturbations of the vapor flux field. Since a certain degree of unrepresentativeness must be expected at almost all island stations, one must be satisfied to eliminate from consideration only those stations that give evidence of being highly biased. Careful investigation of data from each station in this area and initial analyses of flux and flux divergence fields indicated considerable bias in the data from Kingston, Jamaica, under certain flow regimes. The values of flux at the station appeared to be strongly affected by the mountains to the north and northeast of the station, particularly during winter when the prevailing flow is from the northeast. Consequently, little weight was attached to these data. For lack of persuasive evidence to the contrary, all other data were taken at face value. Data were from the 2-yr period May 1961–April 1963.

It was necessary to exclude from consideration the extreme western portion of the Caribbean Sea, since it lies outside the ring of aerological stations. The southwestern and western boundary of the Gulf of Mexico was particularly difficult to handle because of the strong gradients in the flux components and the relatively long distance between stations. In addition, only Merida of the Mexican stations had twice-daily observations, and no data at all were available from Vera Cruz prior to June 1962.

The boundaries used for the computations shown in figure 1 were chosen so that grid point values previously tabulated for use in constructing maps of divergence could be used for these computations as well. Estimation of the boundary flux from a 2.5° grid was deemed adequate in the light of the other uncertainties involved in the compu-

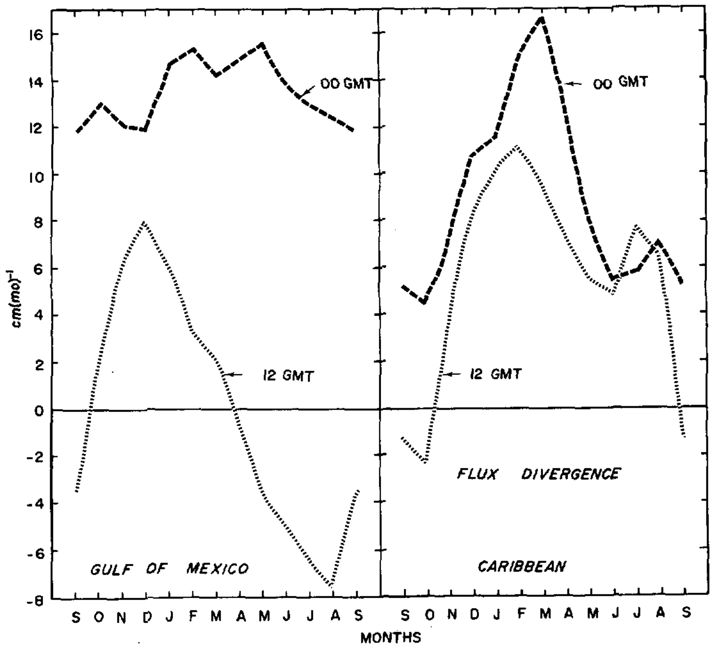

FIGURE 14.—Computed vapor flux divergence at 0000 and 1200 GMT over the Gulf of Mexico and Caribbean Sea.

tation. No corrections were applied for the land areas within the boundary, as it was felt that the station data alone were not sufficiently dense to allow such a distinction to be made. This may not be entirely true near the northern boundary of the Gulf of Mexico.

As pointed out in R2 and illustrated in figure 14, the presence of diurnal variations in the flux divergence field over the Central American Sea and particularly over the Gulf of Mexico gives rise to additional problems in the evaluation of this quantity. Throughout most of the year, the difference between the average divergence over the Gulf of Mexico at 0000 and 1200 GMT is pronounced; and one must recognize the possibility of a significant difference between the true mean value and the average of the 0000 and 1200 GMT observations.

Estimates of the net water transport into these basins by ocean currents are not remotely comparable in accuracy to the runoff measurements over the continent. Consequently, no technique for removing systematic error comparable to that used over land is available.

When recognizing the difficulties involved in the computation of the vapor flux divergence over this area, the results of this study are still believed to be worthy of consideration on a par with values of $\langle\overline{E-P}\rangle$ obtained by estimating $\langle\overline{E}\rangle$ and $\langle\overline{P}\rangle$ individually.

Some correction for atmospheric storage change should be made in spring and fall to avoid slightly biased values of $\langle\overline{E-P}\rangle$ arising because of normal seasonal changes in $\langle\overline{W}\rangle$. For this purpose, mean monthly storage changes were estimated from the 2 yr of data.

CARIBBEAN SEA

Average annual and semiannual values of $\langle\overline{E-P}\rangle$ computed for the Caribbean Sea are shown in table 9, together with three independent estimates of this quantity. Following Wüst (1964), the year is divided into a wet summer season (June–November) and a dry season (December–May).

TABLE 9.—*Water balance components for the Caribbean Sea*

Source	Summer (June–Nov.) $\langle\overline{E-P}\rangle$ (cm 6 mo⁻¹)	Winter (Dec.–May) $\langle\overline{E-P}\rangle$ (cm 6 mo⁻¹)	Annual $\langle\overline{E}\rangle$ (cm yr⁻¹)	Annual $\langle\overline{P}\rangle$ (cm yr⁻¹)	Annual $\langle\overline{E-P}\rangle$ (cm yr⁻¹)
Present study	25	66			91
Colón-Möller (Wüst 1964)	26	63	161	72	89
Budyko (1963)			177	85	92
Hastenrath (1966)	−7	80	199	126	73

Wüst preferred the Colón-Möller estimate over a number of others available at the time of his publication. It is derived from mean monthly estimates of evaporation by Colón (1963) and precipitation charts by Möller (1951); it applies to an elliptic area covering most of the Caribbean and corresponds roughly to the area used for our atmospheric water balance computation. The evaporation figures were obtained from a heat balance computation for the Caribbean Sea, in which the heat flux to the atmosphere was computed as a residual and the flux of sensible and latent heat was separated by assuming a Bowen ratio of 0.10. In his estimates of precipitation, Möller reduced the values of coastal stations by around 20 percent when extrapolating to conditions over the open sea.

The Budyko (1963) estimate is based on charts from his revised "Atlas of the Heat Balance of the Earth." Precise values cannot be determined since they must be estimated by interpolation from isolines. It should be emphasized that this evaporation estimate is obtained from Budyko's revised atlas and is around 30 to 40 cm yr^{-1} higher than the value given in the previous edition. It is the older estimate that is quoted by Malkus (1962), Colón (1963), and Wüst (1964). Note that the Colón-Möller and Budyko estimates of $\langle\overline{E-P}\rangle$ are in better agreement than their estimates of $\langle\overline{E}\rangle$ and $\langle\overline{P}\rangle$ individually, the higher estimates of precipitation used by Budyko being offset by his higher estimates of evaporation.

Hastenrath's values of $\langle\overline{E-P}\rangle$ were obtained from a computation of the vapor flux divergence for the single year 1960. His method of computation differed from that of the present study in that he computed the flux through the boundary of a polygon with stations at the vertices, rather than obtaining values from analyzed maps. The boundaries of the two areas are not very different, with the most significant difference arising in connection with the treatment of the station at Kingston, Jamaica. This station was largely ignored in our analyses, while it served as an important station on Hastenrath's Caribbean perimeter.

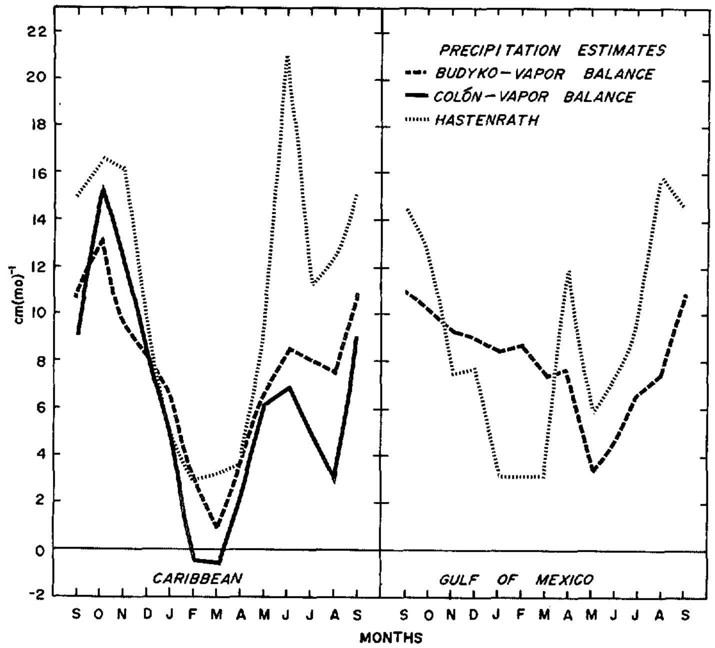

FIGURE 15.—Precipitation estimates for the Gulf of Mexico and the Caribbean Sea.

There is extremely close agreement between annual values of $\langle\overline{E-P}\rangle$ computed from our data and from the estimates of Colón-Möller and Budyko, while those of Hastenrath are 15–20 cm yr^{-1} less. Values for the 2 individual years used in our study were 85 and 96 cm yr^{-1}.

Seasonal values of $\langle\overline{E-P}\rangle$ obtained in this study and the estimates used by Wüst are also in excellent agreement. Hastenrath's values depart sharply, particularly during the summer months.

Figure 15 shows the estimates of mean monthly precipitation that are obtained using our mean monthly values of $\langle\overline{E-P}\rangle$ (smoothed 0.25, 0.50, 0.25) and the evaporation estimates of Colón and Budyko. Also shown are Hastenrath's mean monthly estimates of $\langle\overline{P}\rangle$ for 1960, obtained from land-based stations. The data give seasonal variations in precipitation which are more than twice that of evaporation. The major maximum is computed in October, with a minor maximum in June. The primary minimum is in February or March, with a weak secondary minimum in July or August. For 1960, Hastenrath computed an October maximum of 16.5 cm, close to that obtained from our estimates; but he computed the most pronounced maximum in June (21 cm). Comparison of precipitation data from a number of Caribbean stations for the months of June 1960, 1961, and 1962 indicate that precipitation during June 1960 may have been much greater than the average for the 2 yr used in our study.

GULF OF MEXICO

Computed mean annual values of $\langle\overline{E-P}\rangle$, $\langle\overline{E}\rangle$, and $\langle\overline{P}\rangle$ for the Gulf of Mexico are given in table 10. Values of $\langle\overline{E-P}\rangle$ given by Budyko and those obtained in this study are again in good agreement, while the value computed by Hastenrath for 1960 is around 30 cm less. Values for the 2 yr of our study were 66 and 99 cm yr^{-1}, respectively, suggesting the probability of significant year-to-year variations in the flux divergence. No data from the southwestern Gulf of Mexico were available to Hastenrath during 1960, necessitating a long linear interpolation between Merida, on the Yucatan Peninsula, and Brownsville, Tex. Data from Tacubaya and data from Vera Cruz that became available during the latter part of the period resulted in some improvement during the 2-yr period of our study, although the situation was still far from satisfactory. Hastenrath's estimate of $\langle\overline{P}\rangle$ is fairly close to the climatological value given by Budyko.

TABLE 10.—*Mean annual values (cm yr^{-1}) of the water balance components in the Gulf of Mexico*

Source	$\langle\overline{E}\rangle$	$\langle\overline{P}\rangle$	$\langle\overline{E-P}\rangle$
Present study			82
Budyko (1963)	176	92	84
Hastenrath (1966)	155	99	56

Figure 15 shows the estimate of mean monthly $\langle\overline{P}\rangle$ computed from our smoothed values of $\langle\overline{E-P}\rangle$ and Budyko's values of $\langle\overline{E}\rangle$ and also Hastenrath's values of $\langle\overline{P}\rangle$ for 1960. Both curves show a maximum in late summer-early fall and a minimum in May, but they differ significantly during the winter and early spring.

5. INTERANNUAL VARIATIONS AND INTERREGIONAL RELATIONSHIPS

Mean hydrologic conditions during the period of study have been discussed in previous sections of this paper. Some of the more interesting features of the interannual variability of these data will now be considered. Our attention will be concentrated on the Eastern Region, since the latter part of the 5-yr period marked the early stages of the severe drought that occurred over southeastern Canada and the northeastern United States in the early and mid-1960s.

The onset of the drought is rather well illustrated on figure 16 that shows the cumulative departure from the 4-yr monthly means of precipitation and computed storage for the Truncated Eastern Region. The cumulative departure in Great Lakes storage is also shown. Since the curves represent departures from the mean values for the individual months, the effect of the mean seasonal variation has been removed. Upward- and downward-sloping portions of the curves delineate periods when conditions are respectively above and below the 4-yr normal. A wet period was observed from mid-1959 to mid-1960, while the last half of 1960 and the period beginning in April 1962 mark extended dry spells.

The existence of a high correlation between regional averages of precipitation and storage departures, which was pointed out in R2, is well illustrated in the figure by the correspondence between the precipitation and storage departure curves. Only during the summer and fall of 1961 is there a significant difference in the behavior of the two curves. In this connection, it would probably be well to add a few comments concerning the character of

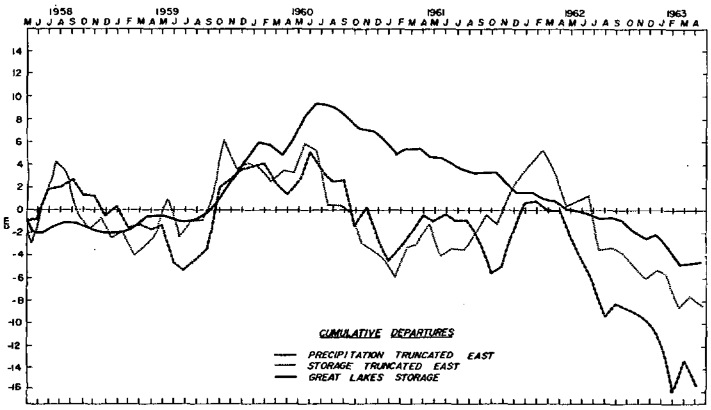

FIGURE 16.—Cumulative departures from the monthly mean values for the 4-yr period May 1958–April 1962.

the departures during the period September 1961–February 1962. During the period of Sept. 10–15, 1961, a huge volume of rainfall fell over an area extending from Texas northeastward across the upper Mississippi Basin into the upper Great Lakes as the remnants of hurricane Carla moved northward from the Gulf of Mexico (La Rue and Younkin 1963). The flux divergence associated with this unusual storm appears to have been reasonably well computed in terms of averages over the combined Central Plains and Eastern Regions, but the distribution between the two regions was not well handled. Generally, the convergence pattern was displaced too far eastward. Problems involved in the computations during this period of intense vapor flux convergence have been discussed by Feruzza (1967). Suffice to say that a more elaborate procedure than that used in this study, one which includes provision for the interpolation of missing data, is called for during such periods. It is estimated that the computed flux divergence over the Eastern Region during this month was probably 2–3 cm too low, thus giving rise to a computed spurious increase in storage. Barring compensating errors prior to February 1961, the actual storage peak at the end of February should be lower and more comparable in magnitude to the precipitation peak.

Great Lakes' levels reached a peak during the summer of 1960, then declined steadily for the remainder of the period. Precipitation departure over the Great Lakes Basin (not shown here) closely followed the trend in lake storage, with an amplitude roughly twice that of the lake storage curve. It is interesting to note that departures of average storage over the Truncated Eastern Region during this 5-yr period were of the same order of magnitude as the average storage change in the Great Lakes. In addition, long-term storage variations over the Truncated Eastern Region were of the same order of magnitude as the seasonal changes.

Flux and flux divergence analyses for the Gulf of Mexico and Caribbean Sea were available for only the last 2 yr of the period. Fortunately, these were years of great contrast over the Eastern Region and the Gulf of Mexico. Figures 17 and 18 show the year-to-year differences in the mean annual values of $\overline{Q}_\lambda$ and $\overline{Q}_\phi$. A decrease in the mean northward and mean westward flux from the first to the second year was observed over the entire Caribbean Sea and over all but the western and extreme northern Gulf of Mexico. Interannual differences in the central Caribbean amounted to more than 400 g(cm s)$^{-1}$, as much as 30 percent of the 2-yr mean at some points The interannual variations of a 10-station average annual mean over a 5-yr period are given in table 11. Interannual variations of more than 30 percent of the 10-station 5-yr mean were observed. Interannual differences at individual stations ranged as high as 950 g(cm s)$^{-1}$. Thus, the year-to-year variability of vapor flux is quite significant over these tropical areas.

Over Eastern North America, except Florida, a decrease in the eastward flux and extremely pronounced decrease in the northward flux from the first to the second year is observed. This was accompanied by a general decrease of 0.10–0.20 g cm^{-2} in the mean annual values of $\overline{W}$ during the second year. In summary, a comparison of mean and departure maps shows the magnitude of the vapor flux vector over Eastern North America and the Central American Sea to have decreased generally from the first to the second year, the major exception being the extreme western Gulf of Mexico. This change in the character of the flux field was accompanied by a decrease in $\overline{W}$ and a sharp decrease in precipitation over Eastern North America.

For further investigation of the relationship between precipitation variations over the Eastern Region and conditions over the Central American Sea, correlations were computed between $\langle\overline{P}\rangle$ for the Eastern Region and the following mean monthly flux components:

1. The meridional flux component at Burrwood, La. This station was taken to represent variations in the northward flux from the Gulf of Mexico.

2. The zonal flux component at Cape Hatteras. This station was taken to represent the eastward flux from Eastern North America.

3. Grand Cayman Island and Swan Island zonal flux components. These stations were taken to represent the westward flux in the Cayman Sea. Correlations between these two stations were also computed to gain some idea of short distance spatial correlations in tropical regions.

4. Curacao zonal component, taken to represent the strength of the westward current in the southern Caribbean Sea. It should be noted here that essentially the same results are obtained if Guadeloupe is used in place of Curacao.

It would be instructive to examine first the correlations between the flux components themselves. (See table 12.) Consider first winter conditions. Correlations in excess of ± 0.60 exist between Hatteras and Swan Island, Grand Cayman, and Burrwood and also between Burrwood and both Swan Island and Grand Cayman Island. The correlation between Curacao and Hatteras is only -0.46, and between Curacao and Burrwood, only -0.10. Thus, variations in the flux over Eastern North America can be traced back through the Gulf of Mexico reasonably well; but their relationship to changes over the Caribbean is, at best, very weak.

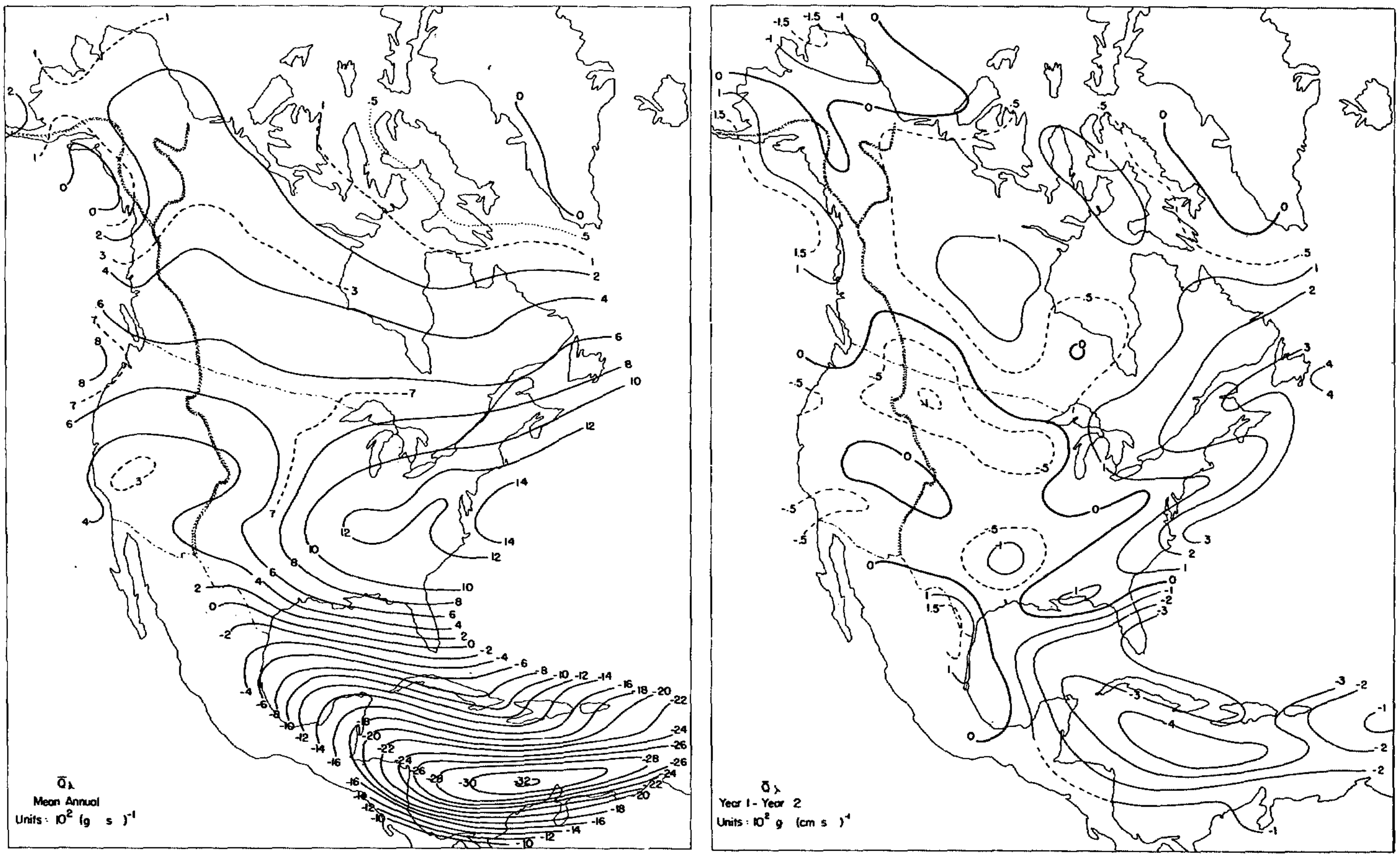

FIGURE 17.—Mean annual values and year-to-year differences of the vertically integrated zonal water vapor flux for May 1961–April 1963; units, 10^2 g(cm s)$^{-1}$. The mean annual chart is from Rasmusson (1967).

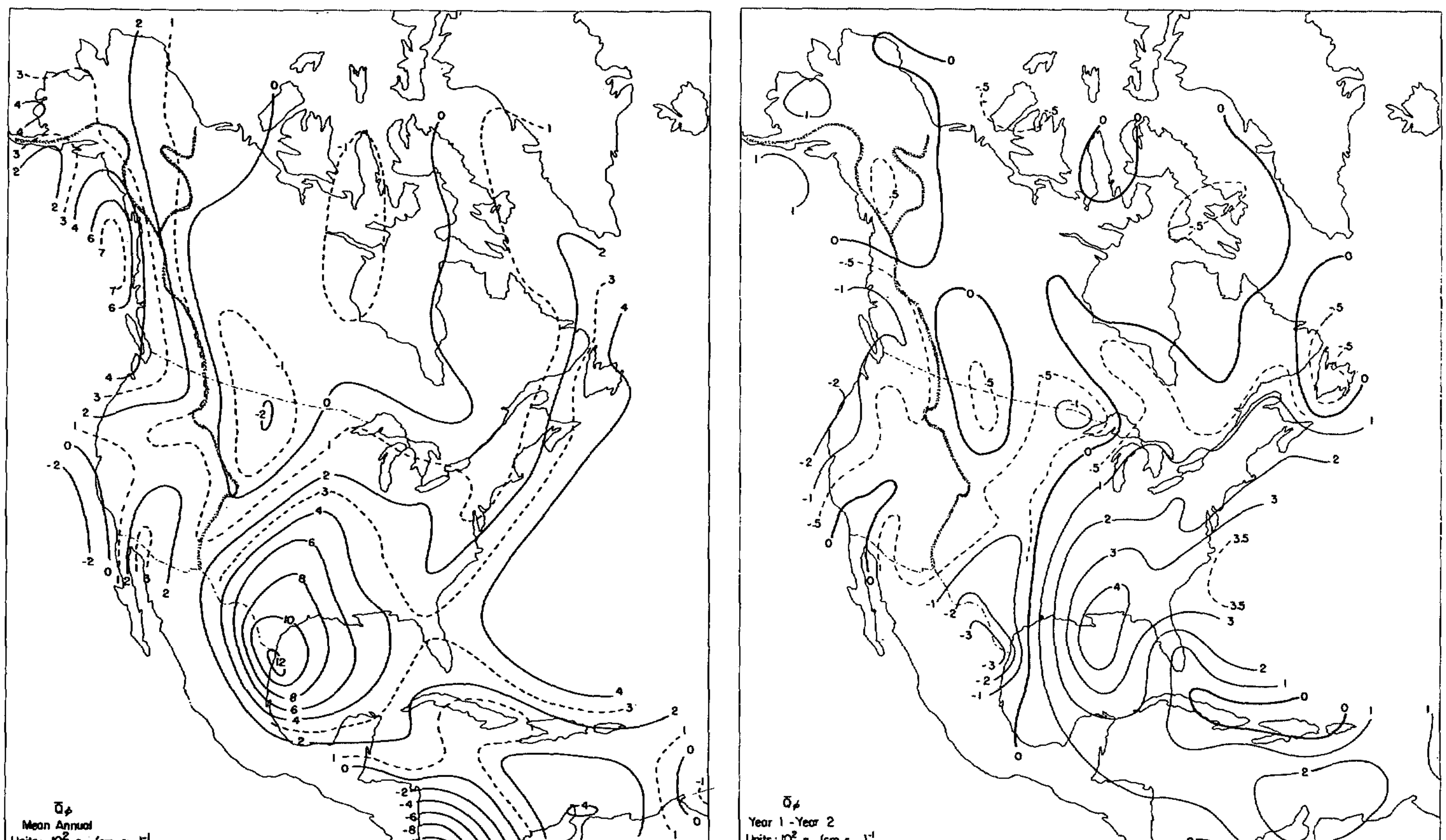

FIGURE 18.—Mean annual values and year-to-year differences of the vertically integrated meridional water vapor flux for May 1961–April 1963; units, 10^2 g(cm s)$^{-1}$. The mean annual chart is from Rasmusson (1967).

TABLE 11.—*Mean annual values of the vertically integrated water vapor flux averaged for 10 stations in the Caribbean-southern Gulf of Mexico area; units, g(cm s)$^{-1}$*

Component	Year (May-April)					
	1958-59	1959-60	1960-61	1961-62	1962-63	5-yr mean
Q_λ	−2355	−2073	−2053	−2101	−1844	−2085
Q_ϕ	145	131	212	69	−289	54

TABLE 12.—*Correlations between mean monthly values of vapor flux (from one-a-day observations during May 1958–April 1963)*

	Winter (Oct.-Mar.)					Summer (Apr.-Sept.)				
	Curacao (zonal)	Swan Island (zonal)	Grand Cayman (zonal)	Burrwood (meridional)	Hatteras (zonal)	Curacao (zonal)	Swan Island (zonal)	Grand Cayman (zonal)	Burrwood (meridional)	Hatteras (zonal)
Curacao	1.00	0.53	0.58	−0.10	−0.46	1.00	0.53	0.45	−0.29	−0.21
Swan Island		1.00	.95	−.65	−.74		1.00	.67	−.39	−.29
Grand Cayman			1.00	−.62	−.72			1.00	−.37	−.39
Burrwood				1.00	.70				1.00	.02
Hatteras					1.00					1.00

TABLE 13.—*Correlations between mean monthly vapor flux and precipitation (from one-a-day observations May 1958–April 1963)*

	Winter (Oct.-Mar.)			Summer (Apr.-Sept.)		
	Burrwood (meridional)	Hatteras (zonal)	Swan Island (zonal)	Burrwood (meridional)	Hatteras (zonal)	Swan Island (zonal)
Eastern Region	0.80	0.66	−0.53	0.03	0.36	−0.18
Great Lakes Basin	.61	.69	−.53	.27	−.05	−.04
Ohio Basin	.80	.57	−.53	.10	.60	−.35

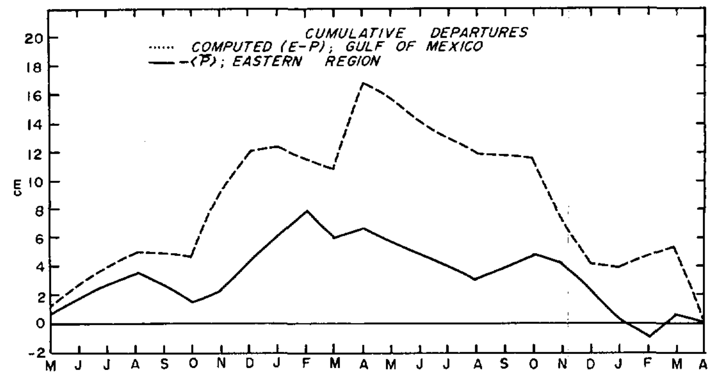

FIGURE 19.—Cumulative departures from the monthly mean values for the 2-yr period May 1961–April 1963.

TABLE 14.—*Comparison of computed net volume of moisture added to the atmosphere over the Gulf of Mexico with precipitation volume over Eastern North America; units, 10^{16} g mo^{-1}*

Region	September	October	November	December	January	February	March	April	May	June	July	August	Annual
E−P Gulf of Mexico	6.0	10.9	13.9	16.0	16.9	16.0	13.5	12.0	10.7	8.1	5.2	3.9	133
P Eastern	18.7	16.7	16.1	16.1	16.7	19.5	17.6	18.9	19.6	22.9	24.8	21.5	229
P Central Plains and Eastern	46.4	36.1	30.5	28.8	26.5	31.8	33.6	36.7	51.6	56.3	59.2	46.9	484

The correlation between Swan Island and Grand Cayman during winter is 0.95, suggesting that a station spacing of roughly 3° would be quite adequate for definition of variations in the mean monthly flux field. The spatial relationship between flux variations is significantly weaker during summer when the correlation between these two stations drops to 0.67, indicating that less than 50 percent of the variance in mean monthly values at a point can be explained by data 350 km away. Summertime relationships between the variations at Hatteras, Burrwood, and Grand Cayman-Swan Island are at best very weak.

Correlations between precipitation departures over Eastern North America and the flux components at Hatteras, Burrwood, and Swan Island are given in table 13. A strong wintertime relationship exists between positive precipitation departures and increased mean northward flux across the Gulf Coast, but again no relationship is apparent during summer.

Finally, the relationship between computed flux divergence from the Gulf of Mexico and precipitation departure over Eastern North America was investigated. Since the flux divergence over the gulf was computed only during the last 2 yr of the 5-yr period, this investigation was of limited scope. Figure 19 shows the cumulative departure from the 2-yr monthly mean values of $\langle \overline{E-P} \rangle$ over the Gulf of Mexico and $-\langle \overline{P} \rangle$ over the Eastern Region.[4] Table 14 gives computed mean monthly values of the total volume of water involved. A negative correlation between the departures apparently existed during this 2-yr period, in that precipitation over the Eastern Region tended to vary inversely with $\langle \overline{E-P} \rangle$ over the gulf, that is, with the input of moisture to the atmosphere from the gulf. This relationship appeared to hold throughout the year, although analysis of a longer period of record is required to firmly establish this fact. It is also apparent

[4] To avoid confusion in the following discussion, note that the negative of the *P* departure has been plotted on figure 19.

from a comparison of the annual computed mean values of $E-P$ over the gulf (66 and 99 cm) with the mean annual variations in the northward moisture flux across the U.S. Gulf Coast, as shown on figure 18, that these quantities also varied inversely from the first to the second year.

The variations in both the mean and eddy flux components will have to be investigated before a clear understanding of these relationships is possible. However, if only winter conditions are considered, the positive correlation between precipitation over the Eastern Region and northward flow of moisture from the Gulf of Mexico is not unexpected since precipitation is generally high during periods when warm moist flow from the south dominates and low during periods dominated by northwesterly flow of cold dry air. It is also not surprising that $\langle\overline{E-P}\rangle$ over the gulf would be highest during periods dominated by a southward flow of cold dry air across the Gulf Coast and lowest during periods when the gulf is dominated by warm moist air masses.

6. FINAL COMMENTS

Water balance computations for areas of Eastern North America varying in size from 42×10^5 km^2 to approximately 5×10^5 km^2 have been presented and discussed. Mean monthly values for the larger areas are consistent and appear to be quite reliable. Results for the smallest areas are sometimes seriously in error.

The computation of interannual storage changes over the Eastern Region strongly indicates that this quantity can be reliably computed for regions of comparable size, using only streamflow data and values of atmospheric flux divergence. The onset of the drought of the early and mid-1960s is clearly reflected in the computed storage values. These results, which may be the most significant to emerge from the budget computations, indicate that interannual storage changes averaged over this area are of the same order of magnitude as seasonal changes.

Interregional relationships between various hydrologic parameters offer an interesting area of inquiry. Results from a few simple computations along these lines show that variations in mean monthly precipitation over the Eastern Region during winter are correlated with the strength of the northward flux of water vapor across the Gulf Coast. Little or no relationship between these quantities exists during summer. In addition, precipitation over the Eastern Region and northward transport of moisture across the Gulf Coast appear to be negatively correlated with the flux divergence over the Gulf of Mexico.

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